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SUMMARY

The evolution and dynamics of the major stratospheric warming of 1984/85 are described and discussed using synoptic maps of geopotential height and temperature, fields of $\underline{Q}$ vectors, isentropic maps of Ertel's potential vorticity, and diagnostics based on zonal averaging and departures therefrom. It is argued that the stratosphere can not be regarded as being pre-conditioned before the warming, in the conventional sense of the term, and that an important element of the warming was that the flow was highly asymmetric at the onset of the rapid breakdown of stratospheric westerlies. Nonlinear dynamics was involved as evidenced by an anticorrelation in time between the amplitudes of zonal waves 1 and 2, by the magnitude of a nonlinear term compared with those of linear terms in a governing equation, and by the irreversible rearrangement of potential vorticity. Apparent loss of potential vorticity in the stratosphere is associated in part with differential advection of air in a vertically sheared flow. Salient features of the large-scale flow in the troposphere during the stratospheric warming are noted.

1. INTRODUCTION

During winter 1984/85 there was a major warming in the stratosphere of the Northern Hemisphere. In this spectacular event, the westerly vortex that characterizes the stratospheric circulation during winter was split in two in a matter of days and temperatures rose by as much as 70°C in some places at 10mb.

In this paper, we describe the warming using observations made by a Stratospheric Sounding Unit on board a NOAA satellite. We focus on the evolution of the stratospheric and tropospheric circulations in the period leading up to the warming, highlighting the major developments and paying particular attention to the vertical structure in the stratosphere and to local changes in the flow.

It has been suggested that for a major warming to happen the stratosphere must be 'pre-conditioned'. Kanzawa (1980) argued that when zonal-mean, westerly winds were confined to high latitudes a polar warming would be favoured because wave activity propagating upwards from the troposphere would be focused into high latitudes. McIntyre and Palmer (1983) extended this idea by considering wave propagation on the main westerly vortex of the stratosphere rather than on its current zonal-mean state. They predicted that the westerly vortex prior to a major warming would be unusually small and therefore more liable to disruption by planetary waves. The major warming of 1984/85 is of note because the stratosphere was not, it seems, pre-conditioned in either of these two ways.

Another intriguing aspect of the warming was that, at its peak, there was a sudden, apparent loss of Ertel's potential vorticity, P, in

our analyses of satellite data on isentropic surfaces. Until then, values of P were relatively well conserved on our maps over periods of about a week. We ascribe this lack of conservation to the formation in the stratosphere of temperature gradients unresolved by our analysis.

Temperatures rose from -60° C to above 0°C locally in the stratosphere during the warming. Such a temperature change cannot be produced by quasi-horizontal advection of temperature by the large-scale flow; vertical motion and attendant adiabatic warming of air must be important. We use so-called $\underline{Q}$ vectors, introduced by Hoskins et al. (1978), to infer where air is rising or sinking with respect to the large-scale flow, and to predict where temperature gradients will increase in the horizontal following the motion.

As part of our study of the warming, we examine the evolution of zonal-mean and eddy components of the flow. Variations are displayed which are characteristic of other warming events. We interpret these variations in terms of localised developments on synoptic maps, and argue that it is inappropriate to treat the dynamics of warmings in terms of waves propagating quasi-linearly on a zonal-mean flow when the flow is as far removed from axial symmetry as it is found to be. A critical assessment is made of current ideas on pre-conditioning, noting carefully the conditions that pertained in the stratosphere and upper troposphere in mid December 1984 at the onset of rapid changes in the circulation.

The paper is laid out as follows. In section 2, we describe the data used in this study and outline how it was processed. Our synoptic account of the warming follows in section 3, where we include a discussion of the behaviour of $\underline{Q}$ vectors and Ertel's potential vorticity

in the mid stratosphere. The evolution of zonal-mean winds and waves is described in section 4 and dynamical implications are considered. We also address the question of pre-conditioning in section 4. Our conclusions are in section 5.

2. OBSERVATIONAL DATA AND METHOD OF ANALYSIS

Our study of the warming of winter 1984/85 is based on daily analyses of data from a stratospheric sounding unit (SSU) on board the satellite NOAA-7. The SSU normally makes radiometric measurements in three channels with weighting functions centred at 15, 5, 1.5 mb - channels 25, 26, 27 respectively - though during the period of present interest channel 26 was inoperative. Although radiances from the SSU have the biggest impact on our stratospheric analyses, measurements from a microwave sounding unit (MSU) and a high resolution, infrared sounder (HIRS-2) on board the satellite are also used. Each instrument is a nadir sounding radiometer which scans across the orbital path. Further details are given by Pick and Brownscombe (1981).

A regression analysis is used to compute thicknesses from the measurements of radiance, and geopotential heights are obtained at standard pressure levels by adding the thicknesses to an analysis of geopotential height at 100mb made daily by the Central Forecasting Office (CFO) of the Meteorological Office. Temperatures are calculated using a finite-difference approximation to the hydrostatic equation. The method used to calculate maps of Ertel's potential vorticity is described by Clough et al. (1985).

The resolution of the measurements is normally about wavenumber 12

in the zonal and meridional (pole to pole) directions, and 12km in the vertical. The loss of channel 26 degrades the resolution of the measurements by the instrument on NOAA-7. To assess the impact on the analyses, we compared them with those made from measurements by an SSU on another satellite, NOAA-6. All three channels of the SSU on NOAA-6 were working but the HIRS instrument on this satellite was not. Measurements by both satellites were analysed in the same way (regression coefficients are specific to an instrument), and both sets of thicknesses were tied to the same field of geopotential height at 100mb. Differences between the two sets of analyses are small even at the peak of the warming. Analyses from NOAA-7 were chosen for this study because more daily analyses were available than there were from NOAA-6.

3. EVOLUTION OF THE MAJOR WARMING OF 1984/85

(a) Synoptic maps of height and temperature in the stratosphere

A dramatic and unusually early major warming developed at the end of December 1984 in the stratosphere of the northern hemisphere. A summary of its main synoptic features has been given by Labitzke et al. (Beilage, 1985), and by Fairlie and O'Neill (1987).

Fig. 1(a) shows the zonally averaged, geostrophic wind on 20 December 1984. There is a broad westerly jet with winds exceeding 80ms^{-1} in the upper stratosphere. (Comparison of data from two satellites indicates that easterlies are overestimated at low latitudes in Fig.1(a) owing to the inoperative channel on the SSU.) By 2 January 1985, Fig.1(b) shows that westerlies are replaced by strong easterlies in the zonal mean throughout

the stratosphere at high latitudes. Near 1mb, zonal-mean easterlies have speeds greater than $70ms^{-1}$, and locally temperatures are about 20°C in the mid stratosphere according to measurements using lidars made at Haute Provence and at the University of Aberystwyth (Labitzke et al., Beilage, 1985, Fig.7).

An important point to appreciate when considering the dynamics of this major warming - a point we elaborate on in section 4 - is that the flow depicted in Fig.1(a), at the start of rapid deceleration of zonal-mean winds, is far from being axisymmetric about the pole, or about any other axis. Fig.2 shows the fields of geopotential height and temperature at 10mb on 20 December 1984. A large, elongated, cyclonic vortex is displaced from the pole by a strong anticyclone near the dateline, the Aleutian High. The coldest air is associated with the cyclone, and a tongue of relatively warm air extends westwards from the anticyclone.

During the following week, the Aleutian High weakens slightly, but another intense anticyclone develops over the Atlantic and the westerly vortex starts to split. The Atlantic anticyclone develops through a deep layer: the closed circulation associated with it extends from the upper troposphere, where the circulation is part of a 'block', to the stratopause. Fig.3 shows the fields of geopotential height and temperature on 28 December 1984. Temperatures are rising sharply at the entrance to a southerly jet stream that is forming to the west of the Atlantic anticyclone. As temperatures rise, the anticyclone tilts westwards with height. After 28 December, the Aleutian High also intensifies and the westerly vortex splits into two roughly equal parts. The anticyclones merge to form one large anticyclone associated with a pool of warm air over the polar cap, as shown by Fig.4, which shows the fields of geopotential height and temperature on 2

January 1985.

After the westerly vortex splits, the circulation in the lower and middle stratosphere stays disrupted throughout January because (1) the radiative damping time there is quite long, about 20 days (Dickinson, 1973, and others); and (2) the troposphere remains highly asymmetric on a planetary scale. In the upper stratosphere, the damping time is shorter - about 5 days (Dickinson). The westerly circulation there quickly recovers at mid and low latitudes, but not at high latitudes where the Aleutian High extends upwards into the middle of the westerly vortex until early February.

(b) $\underline{Q}$ vectors

Local rises in temperature during the warming far exceed what could be achieved by quasi-horizontal, warm advection. Vertical motions and adiabatic warming must be important, as Blackshear et al. (1987) found in a stratospheric warming simulated by a general circulation model. Vertical velocities are not measured by the SSU radiometer, but the direction of vertical motion may be inferred by using a form of the quasi-geostrophic 'omega equation' derived by Hoskins et al. (1978, Eq.(7)). Hoskins and Pedder (1980) used the equation to diagnose the development of a baroclinic wave at mid-latitudes in the troposphere.

Central to the omega equation is a two-component quantity called the $\underline{Q}$ vector, defined by

$$\underline{Q} = -\left\{ \underline{\nabla}\theta \cdot \partial \underline{V}_g / \partial x \, , \, \underline{\nabla}\theta \cdot \partial \underline{V}_g / \partial y \right\} \qquad (1),$$

using standard notation given in the Appendix. Hoskins and Pedder argue that $\underline{Q}$ should point towards rising motion; upward motion is expected where $\underline{Q}$ converges in the horizontal, and downward motion where it diverges. They write the following equation:

$$\left(\frac{\partial}{\partial t} + \underline{v}_g \cdot \underline{\nabla}\right) |\underline{\nabla}\theta|^2 = 2\,\underline{Q} \cdot \underline{\nabla}\theta \qquad (2).$$

The term $2\underline{Q}\cdot\underline{\nabla}\theta$ is called the geostrophic frontogenesis function. It gives a measure of the rate of change, due to the geostrophic winds, of potential temperature gradients following the geostrophic motion (Hoskins and Pedder).

Fig.5 shows the $\underline{Q}$ vectors at 10mb on 28 December 1984, together with isopleths of potential temperature. Fig.3 is the corresponding map of geopotential height and temperature. Having calculated $\underline{\nabla}\cdot\underline{Q}$, we find that, in the jet stream near 90°W, there is a maximum in divergence at the point marked A in Fig.5 and a maximum in convergence at B. So we expect air to sink as it enters this jet stream and rise as it leaves it. (The vertical motion varies similarly along the jet stream near 90°E.) The warmest air should therefore lie downstream from A and upstream from B, i.e. at the lowest point in an air parcel's trajectory where it has warmed the most adiabatically. This prediction is consistent with Fig.3.

The $\underline{Q}$ vectors point towards high values of θ just east of A, so the frontogenesis function is positive, indicating <u>potential</u> frontogenesis there - a process that <u>might</u> culminate in the formation of a front. There is no such effect at B where $\underline{Q}$ vectors are almost tangential to contours of θ. This diagnosis is in accord with the actual evolution of temperature gradients over the next few days. Gradients increase in the jet stream near 90°W until about 30 December. As the westerly vortex splits, $\underline{Q}$ vectors lie parallel to,

rather than across, θ contours in the jet, and temperature gradients do not increase further.

The configuration of $\underline{Q}$ vectors with respect to streamlines of the large-scale flow is similar during the build-up of the warming to that found by Hoskins and Pedder (1980) in their diagnosis of a growing baroclinic wave in the upper troposphere.

It is likely that maximum temperature gradients are underestimated significantly in the observed fields because of limited resolution. In a reasonably accurate simulation of the warming with a numerical model (grid points spaced at 5° intervals in the horizontal and 2.5 km in the vertical), much stronger temperature gradients are found, and structures are produced resembling fronts in the upper troposphere (Fairlie, Fisher and O'Neill, paper in preparation). Vertical velocities from the simulation broadly confirm the predictions of the $\underline{Q}$-vector analysis.

We conclude that fields of $\underline{Q}$ vectors give useful diagnostic information during stratospheric warmings, allowing the direction of vertical motion to be inferred from large-scale, measured fields, and showing how temperature gradients are tending to change following the motion. The same conclusion was reached by Hoskins and Pedder with regard to baroclinic waves in the troposphere.

(c) Ertel's potential vorticity, P

Isentropic maps of Ertel's potential vorticity, P, allow the movement of material to be followed for short periods during the warming, albeit with a blurred view because of limited resolution. As in Clough et al. (1985), P is computed from the following expression:

$$P = -g(\zeta + f)\,\partial\theta/\partial p \qquad (3),$$

using standard notation given in the Appendix.

Fig.6 is such a map for 28 December 1984 and for the 850K isentropic surface, which lies near 10mb at a height of about 30km in the mid stratosphere. Arrows representing the speed and direction of the horizontal winds are also shown on this surface. The circulation associated with an anticyclone over the Atlantic is drawing out a tongue of high P into low latitudes. The area covered by cyclonic circulation shrinks as air with high P is drained away in this way. Such advection of material by developing anticyclones is common in the winter stratosphere. Its occurrence was first pointed out by McIntyre and Palmer (1983). They regard such irreversible (time-oriented) advection of P as being the hallmark of breaking planetary waves, though, as O'Neill and Pope (1987) remark, the phenomenon is not restricted to cases for which a propagating wave is readily defined.

We are assuming in interpreting Fig.6 that P contours are approximately material lines. Two reasons this assumption gets poorer as the warming develops are: (1) temperatures are much higher than they would be in radiative equilibrium, so air cools more rapidly and P is not conserved for so long; (2) temperatures rise in the upper stratosphere and warm air overrides cold air; the vertical resolution of the SSU is then a limitation.

An illustration of (2) is the poor conservation of P at the height of the warming in our analysed fields. Fig.7 shows the distribution of P and winds on 30 December 1984 when the cyclone has split in two. Maximum values of P ought to be conserved for a few days at least, though advection might move the maxima somewhere else on an isentropic map. Fig.8 for 2 January 1985 shows, however, that peak values are halved for one of the vortices and

almost so for the other. The reason is that there are now sharp changes in vertical temperature gradient, which the SSU cannot resolve.

Fig.9 shows the variation of temperature with height recorded by a radiosonde launched from Berlin on 2 January. Whereas the lower stratosphere is nearly isothermal, there is a highly stable layer in the mid stratosphere where temperatures increase with height at about 10K/km. The dashed curve on the figure is computed from analyses of geopotential height derived using SSU data. The profile is colocated with the Berlin radiosonde. The sharp transition to a stable layer is not captured by the satellite measurements; the stability of the mid stratosphere is underestimated, and so, from Eq.(3), is P. Fig.9 clearly exposes the limitations of measurements from the SSU during strong warmings. But there is mitigation. We find with Clough et al. that P derived from SSU data is usually well conserved from day to day. So when it is not, we are warned that our analysis is poorly representing the state of the stratosphere. This is one reason for studying isentropic maps of P.

With the split in the vortex, an anticyclonic circulation enclosing air drawn from low latitudes (as indicated by low values of P) encompasses the polar cap (Fig.8). The two cyclones are being stretched around it. Because the easterly winds around the anticyclone increase in speed with height, we might expect that the cyclones would move westwards faster in the upper stratosphere than in the middle stratosphere. This is confirmed by our analyses (with the caution that they are not as reliable at this stage of the warming as they are at its inception). We ascribe the loss of resolution in the vertical to a combination of such quasi-horizontal, <u>differential advection</u> of P and adiabatic warming as air sinks.

Strong vertical temperature gradients that are more in keeping with the

observations made by radiosondes and lidars are reproduced in the numerical simulation mentioned above.

(d) Evolution of the large-scale flow in the troposphere

The salient feature of the circulation in the upper troposphere and lower stratosphere in early December 1984 was a 'low' in geopotential height in the sector 90°E to 180°E, as shown in Fig.10, the synoptic map at 100mb for 5 December. At this time in the mid and upper stratosphere, the Aleutian High was intensifying. It can be regarded, in part, as a downstream development from the localised 'low' in the troposphere (O'Neill and Pope, 1987).

Also shown in Fig.10 is a trough over Canada, which subsequently intensifies to form a closed cyclonic circulation. On the synoptic map of geopotential height at 100mb for 20 December, Fig.11, there are two cyclones separated by two ridges; the pattern now contains a distinct wave-2 component. By contrast, the flow earlier in the month was dominated by wave 1 (Fig.10).

The middle and upper stratosphere participates in this transition from a wave-1 dominated to a wave-2 dominated flow. Cyclogenesis in the troposphere over Canada precedes the formation downstream of a closed anticyclonic circulation in the stratosphere over the Atlantic (Figs.3 and 6). This growing anticyclone, in concert with the Aleutian High, eventually splits the stratospheric cyclone.

4. RELEVANCE OF CONCEPTS OF WAVE PROPAGATION AND PRE-CONDITIONING TO THE MAJOR WARMING OF 1984/85

Much of the dynamical thinking about stratospheric warmings starts from the premise that planetary waves can be considered as quasi-linear perturbations propagating upwards on some 'basic state' of the stratosphere. The much-studied major warming of February 1979 is a case in point. Palmer's (1981) analysis of the event is based on this assumption, and the series of idealised experiments done by Butchart et al. (1982) was designed to demonstrate that, by controlling the propagation of planetary waves, an anomalous zonal-mean state before the warming was crucial for its later development.

McIntyre (1982) suggested that it would be more relevant, when the stratospheric cyclone was distorted and displaced from the pole, to consider wave propagation not on the zonal-mean state but on a state obtained by making the actual flow symmetric in a way that preserved the areas enclosed by potential vorticity contours. Pursuing this idea, McIntyre and Palmer (1983) predicted that the area of the westerly vortex in the stratosphere would be unusually small prior to a major warming. They reasoned that wave activity would be more focused by a small vortex than by a large one, and so stand a better chance of disrupting it.

The term 'pre-conditioning' has been applied to the process that produces a basic state favourable to a major warming (e.g. Kanzawa, 1980). McIntyre and Palmer (1983) proposed that the mechanism is irreversible rearrangement of potential vorticity during 'planetary wave breaking', which causes to the westerly vortex to shrink.

A basic difficulty with the above conception of the dynamics of major warmings is that prior to them - we shall clarify later what we mean by 'prior' - the flow is highly asymmetric about any axis. Labitzke (1977) refers to the wave-1 pre-cursor, alluding to the presence of an intense Aleutian High in the stratosphere before the rapid breakdown of the westerly circulation. Thus, there is no reason to suppose that it makes dynamical sense to 're-symmetrize' this distorted flow in order to define a basic state on which disturbances can be regarded as propagating quasi-linearly, nor to suppose that the area of the westerly vortex provides enough information to judge whether one asymmetric stratospheric state favours a major warming more than another.

Two dynamical aspects of the warming of 1984/85 that we shall now consider are these:

(1) There is a strong anticorrelation in time in the amplitudes of the principal zonal wavenumbers 1 and 2 throughout the stratosphere and upper troposphere. We shall argue that this indicates a nonlinear interaction between the harmonics, precluding the use of the linear theory of planetary waves.

(2) The stratosphere was not obviously pre-conditioned in the above sense. Before the warming the zonal-mean jet was broad and strong, and the westerly vortex was not unusually small for winter. It was, however, highly distorted.

Consider first the evidence for point (1). Fig.12(a) is a height-time section at 60°N of the amplitude of zonal wave 1 in geopotential height from mid November to mid January. Fig. 12(b) is the corresponding section for wavenumber 2. In the mid and upper stratosphere in early December, a temporary minimum in wavenumber 1 coincides with a maximum in wavenumber 2.

In mid December, the amplitude of wavenumber 1 is more than twice that of wavenumber 2. Then, in the latter half of the month, the circulation undergoes a rapid change, reflected in an abrupt transition from wave 1 to wave 2 almost simultaneously over a deep layer from the upper troposphere to the stratopause. On a sequence of synoptic maps, the nonlinear transition is associated with the development of an anticyclone 180° in longitude away from the Aleutian High. By the end of the month, the amplitude of wave 2 far exceeds that of wave 1. Such an anticorrelation between wavenumbers 1 and 2 in the stratosphere has been noted by many authors (e.g. Hirota and Sato, 1969).

The transition is not, in itself, unequivocal evidence for nonlinear coupling between the two harmonics. Nonlinearity does seem likely though, first because maximum amplitudes of both harmonics are unusually large in the stratosphere in December 1984, and second because contours of P deform irreversibly near the growing anticyclone over the Atlantic (section 3(c)).

To prove that nonlinear dynamics is involved, we have followed O'Neill and Pope (1987) in taking Eq.(4) below as the basic equation governing the evolution of the flow - invoking, as they did, the 'invertibility principle' discussed by Hoskins et al.(1985) - and comparing the linear and nonlinear terms in it.

$$dP/dt = \partial P/\partial t + \underline{V}_0 \cdot \underline{\nabla} P_0 + \underline{V}' \cdot \underline{\nabla} P_0 + \underline{V}_0 \cdot \underline{\nabla} P' + \underline{V}' \cdot \underline{\nabla} P' \qquad (4)$$

$$= \text{non-conservative terms}$$

In this equation, primes denote departures from a reference value denoted by the subscript zero. The reference state is taken to be the flow on 20 December, at the start of the wave-1 to wave-2 transition. If the transition

between harmonics was, in fact, due to linear superposition on the asymmetric reference state of a new disturbance involving both waves 1 and 2 (the wave-1 components of the original and new disturbances being in anti-phase), the term in Eq.(4) involving products of primes would be small compared with terms with only one prime.

We find that by the time wave-2 amplitudes exceed those of wave 1 in the stratosphere, maximum and minimum values of the nonlinear term in Eq.(4) are similar to those of the linear terms throughout the stratosphere. Moreover, the linear and nonlinear terms vary on the same meridional scale. That is, the nonlinear domain is too broad and too deep to use linear wave theory reliably anywhere in the winter hemisphere during the build-up to the major warming. O'Neill and Pope made similar findings in their analysis of a major warming produced in an integration with a numerical model. So predictions about the likelihood of a major warming based solely on considerations of linear wave dynamics cannot be trusted, as we now demonstrate by considering point (2) above.

The structure of the zonal-mean wind prior to the major warming of 1984/85, shown in Fig.1(a), is typical of a climatological zonal-mean wind (shown, for example, by Murgatroyd, 1969): the zonal-mean jet is broad and the jet core in the upper stratosphere is near mid latitudes, not at high latitudes as in so-called pre-conditioned states. (As mentioned in section 3(a), easterlies are probably too strong at low latitudes in Fig.1(a); measurements from the satellite NOAA-6 indicate that the zero-wind line at 10mb is at 10°N.) Since the zonal-mean state is not unusual, it can hardly be regarded as being pre-conditioned.

This conclusion cannot be avoided by supposing that planetary-scale disturbances in the troposphere were unusually strong in December 1984. From

analyses of the troposphere and lower stratosphere made by the National Meteorological Center (Washington) and the Meteorological Office, we find that amplitudes of zonal waves 1 and 2 in geopotential height were indeed large, but not unusually so compared with peak amplitudes in some winters that Labitzke (1982) has classified as having minor warmings but not a major warming.

We have adopted McIntyre and Palmer's (1984) definition of the size of the westerly vortex, and computed this size on the 850K isentropic surface before a number of warmings by following the prescription they give:

(1) Compute the area on the isentropic surface, A(P), where the value of potential vorticity is bigger than P.

(2) Repeat for many values of P and then compute the derivative, dA/dP.

(3) Find the value of P, P_*, at which there is a sharp change in dA/dP.

(4) A(P_*) is the measure we use of the size of the vortex.

To determine whether the westerly vortex was unusually small before the major warming of 1984/85 compared with its size before minor warmings, it is necessary to give an objective definition of what is meant by 'before'. The reason is that, during the course of a warming, the vortex shrinks (McIntyre and Palmer, 1983; Butchart and Remsberg, 1986). This shrinkage is, apparently, as much a characteristic of a stratospheric warming as is a rise in polar temperature or a deceleration of the zonal-mean wind. It therefore seems reasonable to compare areas <u>at the onset</u> of rapid changes in some aspect of the flow. The standard (W.M.O.) classification of warmings as major or minor hinges on the extent to which zonal-mean easterlies replace westerlies in the stratosphere. So we take the onset of the warming to be when rapid deceleration of the zonal-mean wind begins, typically about a week or so before the height of the warming. Such a period is commonly regarded as the

time-scale for a warming to develop. Butchart et al. (1982), for example, initialized their numerical experiments on the warming of February 1979 using observed fields about a week before the vortex split.

In Fig.13 are plotted values of $A(P_*)$ for the 850K surface before major and minor warmings that happened in the northern hemisphere from the end of 1978 to the beginning of 1986. Warmings deemed major by Labitzke et al. (1979 to 1986, Beilage, published annually) are denoted by an asterisk. All warmings satisfying the following criteria are included:

(1) The warmings are strong enough to warrant reference in the annual summaries of the winter circulation issued by Labitzke et al. (op. cit.). In 5 days, zonal-mean winds at 60°N,30mb decelerate by 10 m s^{-1} or more, as deduced from the plots given by Labitzke et al..

(2) The warming falls in the period December-February, i.e. it is a mid-winter warming rather than a final warming - the annual reversal of winds in spring. Final warmings are included if they fall in the above time span.

The figure shows that for about half the minor warmings plotted (points lying below the dashed line) the vortex was smaller than it was for the major warming of 1984/85. Moreover, this major warming is not exceptional in starting from an unusually big vortex: there are two more major warmings plotted for which the vortex was bigger still. McIntyre and Palmer (1983) made their prediction about vortex size on the evidence of the major warming of February 1979 but, if anything, it is this case that is unusual. Indeed, Labitzke (1982) refers to it as an early final warming rather than as a major mid-winter warming. So it seems from our results that there is no predictive power in using $A(P_*)$ to forecast (about a week ahead) that a warming in the northern hemiphere, should one happen, will be major rather

than minor.

These findings do not mean that the size of the vortex has no bearing on the stratospheric response to a given tropospheric disturbance. On the contrary, the idealised experiments of Butchart et al. (those starting with a zonally symmetric flow in the stratosphere and upper troposphere) clearly demonstrate that size does matter. But our results imply that the size of the westerly vortex before a warming in the real northern hemisphere is only one - and apparently not the dominant one - of a number of factors controlling what happens.

O'Neill and Pope (1987) suggest that a highly distorted vortex may be more susceptible to being broken down than a symmetric one. This idea fits with Labitzke's (1977) observation, valid for the warming of 1984/85, that a strong wave 1 is a precursor to major warmings dominated by either zonal wave 1 or wave 2. To illustrate the point, consider the following thought experiment. In a numerical model of the middle atmosphere with pure wave-1 forcing at its lower boundary (near the tropopause), a strong wave 2 is generated nonlinearly in the stratosphere, according to Palmer and Hsu (1983). Now add to the wave-1 forcing a strong perturbation in wave 2. The stratospheric response would depend *inter alia* on the phase of the new wave: the westerly vortex in the stratosphere might split if the applied wave 2 reinforced its nonlinearly generated counterpart. So if, in the real atmosphere, the tropospheric circulation partakes with the stratosphere in developing a strong perturbation in wave 2 after one in wave 1 (because of the constraint imposed by the conservation of P, this would appear as a nonlinear transition), a split vortex and a major warming might result. This seems to be what happened in December 1984 and, incidentally, in the famous warming of February 1979. If the troposphere does not do so - and given the

complexity of its dynamics it may not every time - then the stratospheric westerly vortex is not likely to split but only to be displaced from the pole and elongated.

5. CONCLUSIONS

The main dynamical elements of the major warming of 1984/85 that we identify and discuss are these:

(1) The stratosphere was not pre-conditioned before the warming in the conventional sense of the term: zonal-mean winds were typical of a climatology for northern winters; and the westerly vortex was not unusually small. We were unable to predict from the size of the stratospheric vortex about a week beforehand that a warming in the northern hemisphere would be major rather than minor.

(2) The westerly circulation broke down when an intense anticyclone developed over the Atlantic in a stratosphere already containing a strong Aleutian High and an elongated westerly vortex. These changes were manifested as a nonlinear transition from zonal wave 1 to wave 2, the nonlinearity being significant over much of the stratosphere and upper troposphere.

(3) As far as we could judge diagnostically, cyclogenesis around mid December in the troposphere over Canada and attendant downstream development of an anticyclone in the stratosphere were instrumental in splitting the cyclone in the stratosphere.

(4) Isentropic maps of Ertel's potential vorticity, P, showed that the warming was associated with irreversible rearrangement of P, leading to a shrinkage of the westerly vortex, as envisaged by McIntyre and

Palmer (1983,1984,1985) in their wave breaking scenario.

(5) Rapid apparent loss of high values of P after the westerly vortex split was ascribed in part to differential advection in a vertically sheared flow, which reduces vertical scales below those that could be fully resolved.

(6) So-called Q vectors allowed the direction of vertical motion to be inferred and showed how temperature gradients tend to change following the large-scale flow.

A simulation of the major warming of 1984/85 using a numerical model of the stratosphere and mesosphere based on the primitive equations will be presented in a forthcoming paper (Fairlie, Fisher and O'Neill; in preparation). Features which were simulated by the model but which could not be adequately resolved in the satellite observations were narrow, westward-leaning zones of strong temperature gradients - or 'fronts'.

ACKNOWLEDGMENTS

We thank Jacqueline Lloyd for doing the calculations for Fig.13 and Betty Kingston for preparing most of the diagrams. David Andrews, Victoria Pope and the referees gave us useful comments on the paper.

APPENDIX

$A(P)$ area on isentropic surface where potential vorticity is bigger than P

$A(P_*)$ 'size of the stratospheric westerly vortex'

f Coriolis parameter

g	acceleration due to gravity
p	pressure
P	Ertel's potential vorticity
P_*	value of potential vorticity at which there is a sharp change in dA/dP
$\underline{Q}$	two-component quantity defined by Eq.(1), section 3(b).
$\underline{V}$	horizontal velocity
$\underline{V}_g$	geostrophic wind
x	distance east
y	distance north
t	time
ζ	relative vorticity
$\underline{\nabla}$	horizontal gradient operator
θ	potential temperature
$( )_o$	value of variable in reference state
$( )'$	departure from reference state

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FIGURE LEGENDS

Fig.1. (a) Cross-section of zonal-mean, geostrophic wind for the northern hemisphere on 20 December 1984. Contours are plotted at intervals of 10 ms^{-1}, and shading denotes easterly winds.
(b) The same but for 2 January 1985.

Fig.2. Synoptic map at 10mb of geopotential height (solid lines) and temperature (dashed lines) for 20 December 1984. Units: dam for geopotential height; K for temperature.

Fig.3. As for Fig.2 but for 28 December 1984.

Fig.4. As for Fig.2 but for 2 January 1985.

Fig.5. Synoptic map at 10mb of Q vectors (arrow at bottom left represents a magnitude of 50 K m^{-1} s^{-1}) and potential temperature (K) on 28 December 1984.

Fig.6. Synoptic map of Ertel's potential vorticity, P, and winds on the 850K isentropic surface, near 10mb, for 28 December 1984. Contours of P are plotted at intervals of 10^{-4} K m^2 kg^{-1} s^{-1}. In these units, dark shading denotes values between 5 and 3; light shading denotes values between 2 and 1. Wind speeds can be read using the scale at the bottom left of the figure.

Fig.7. As for Fig.6 but for 30 December 1984.

Fig.8. As for Fig.6 but for 2 January 1985.

Fig.9. Co-located profiles of temperature (°C) on 2 January 1985. The solid curve is plotted from measurements made by a radiosonde launched from Berlin. The dashed curve is derived by differentiating in the vertical fields of geopotential height obtained from SSU data.

Fig.10. Synoptic map of geopotential height (dam) at 100mb on 5 December 1984.

Fig.11. As for Fig.10 but for 20 December 1984.

Fig.12. Height-time section at 60°N of amplitudes (m) of zonal harmonics in geopotential height; (a) wave 1, (b) wave 2.

Fig.13. The sizes, $A(P_*)$, of the westerly vortex, measured on the 850K isentropic surface, before major and minor warmings in the stratosphere of the northern hemisphere that occurred in the months December, January or February (labelled D,J or F on the abscissa) from December 1978 to February 1986. Points marked with an asterisk are for major warmings, other points are for minor warmings. The letter 'C' next to a point denotes a Canadian warming, the letter 'F' a final warming. A point marked by both an asterisk and the letter 'F' (e.g. the one in February 1979) is called a 'major final warming' by Labitzke et al. (1979 to 1986). The dashed line passes through the point for the major warming of 1984/85. See main text for details of warmings included in the diagram and for the method of calculation of $A(P_*)$.

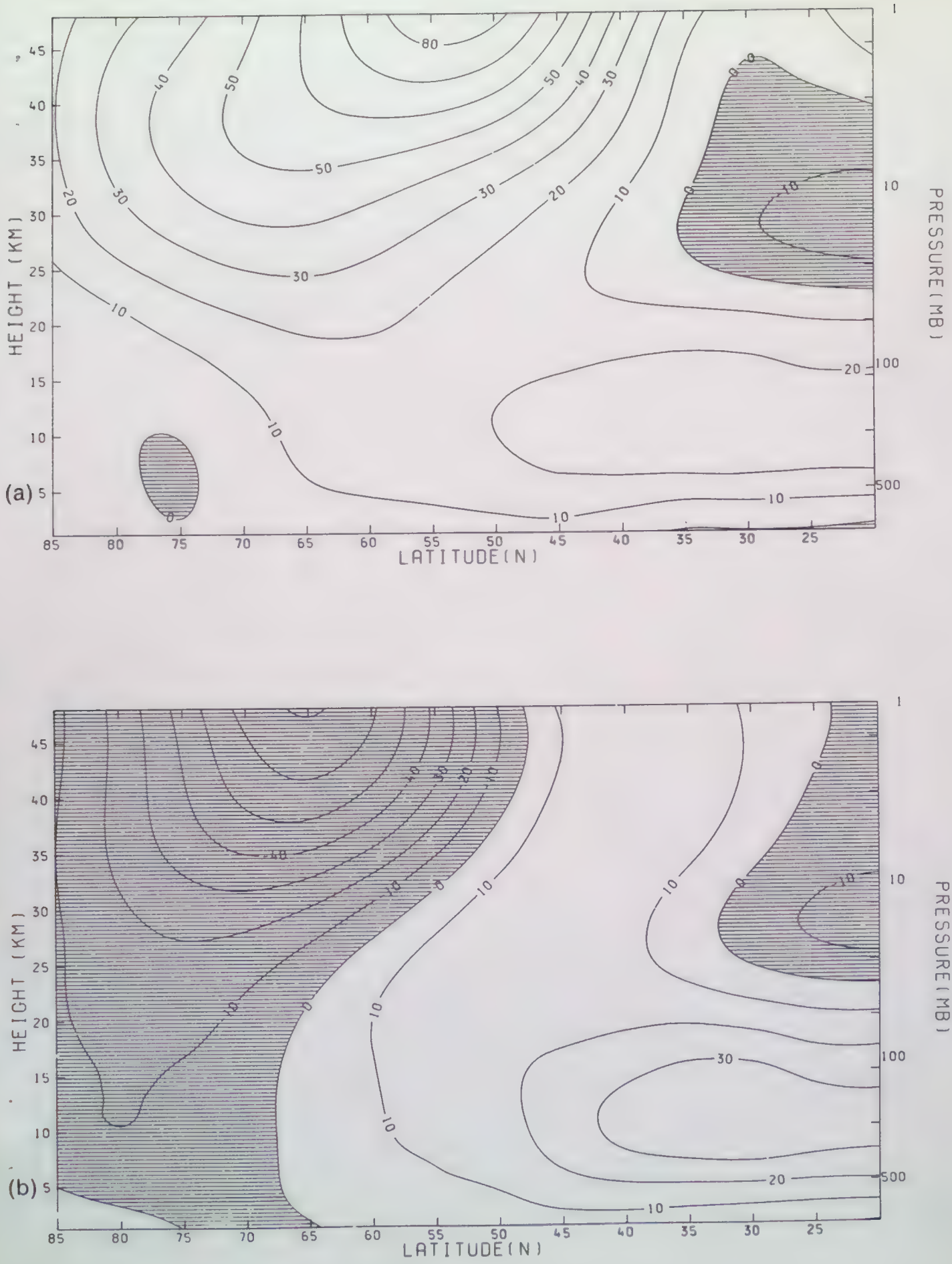

(a)
(b)
HEIGHT (KM)
LATITUDE(N)
PRESSURE(MB)
85 80 75 70 65 60 55 50 45 40 35 30 25
5 10 15 20 25 30 35 40 45
1 10 100 500

Figure 1

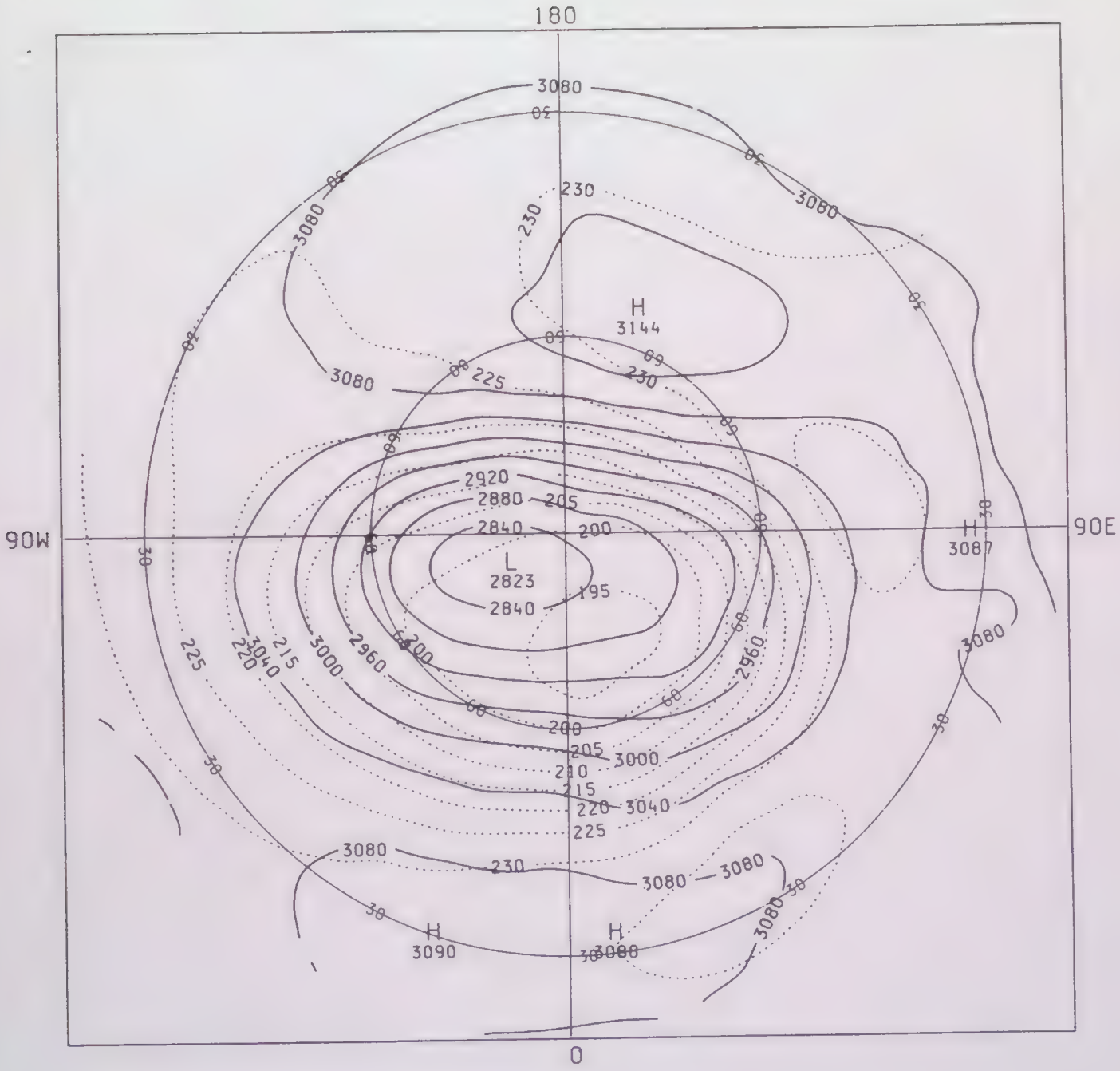

180
90W
90E
0
H
3144
L
2823
H
3087
H
3090
H
3088
3080
3040
3000
2960
2920
2880
2840
230
225
220
215
210
205
200
195

Figure 2

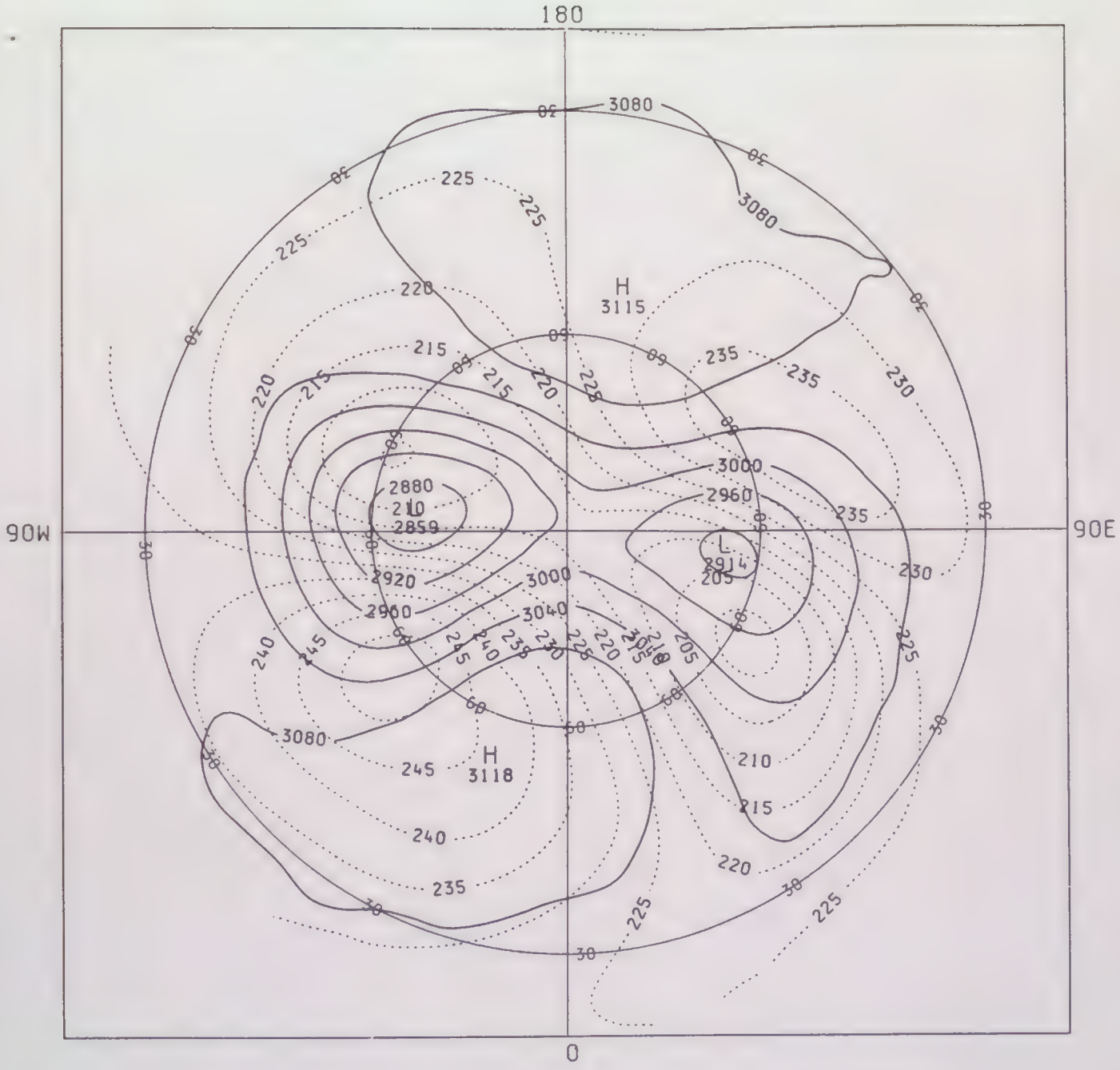

180
90W
90E
0
H
3115
H
3118
L
2914
3080
3000
2960
2880
2859
2920
2900
3040
225
220
215
235
230
240
245
210
205
30
60

Figure 3

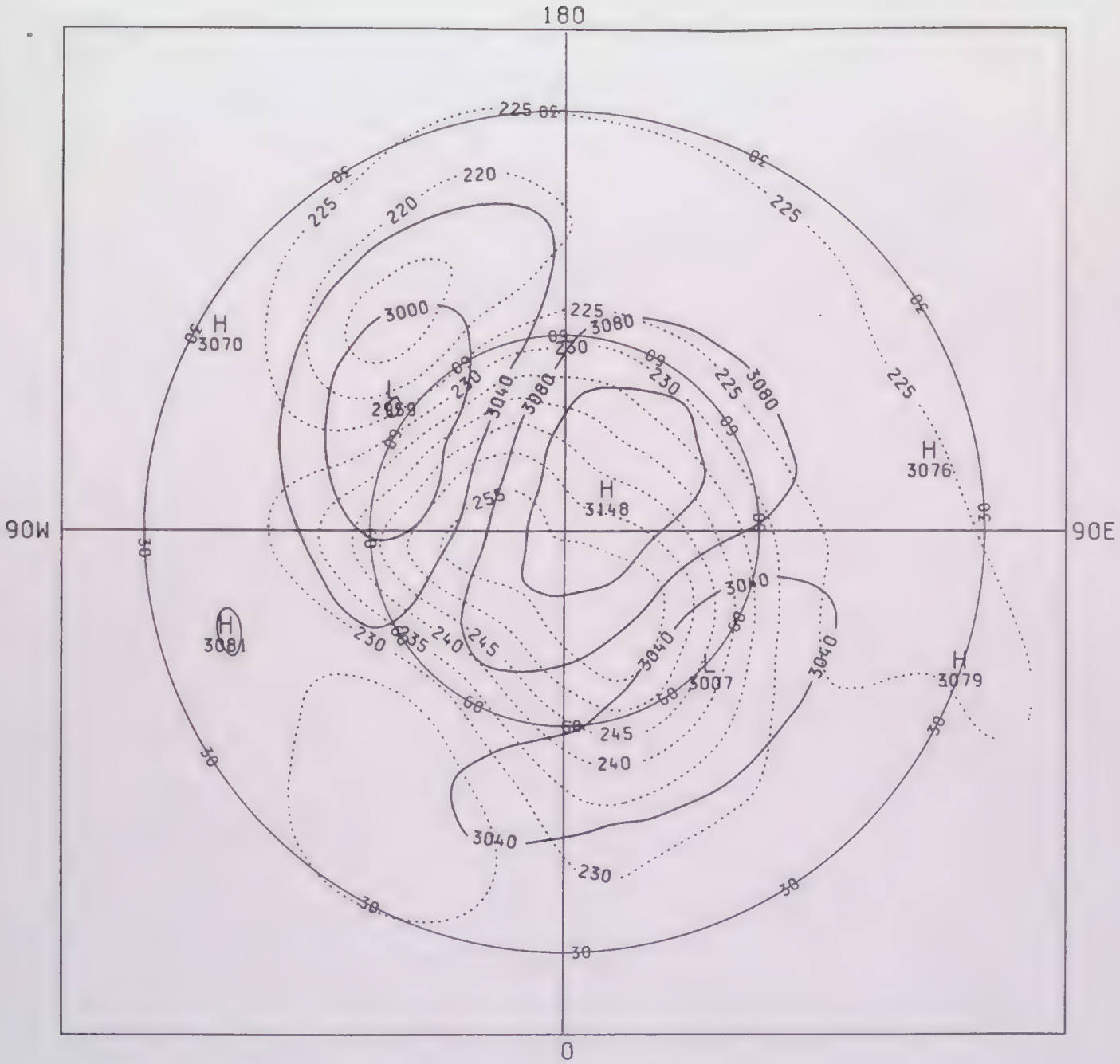

180
90W
90E
0
H
3070
H
3076
H
3148
H
3081
H
3079
L
3007
3000
3040
3080
225
220
230
240
245
255
30
60

Figure 4

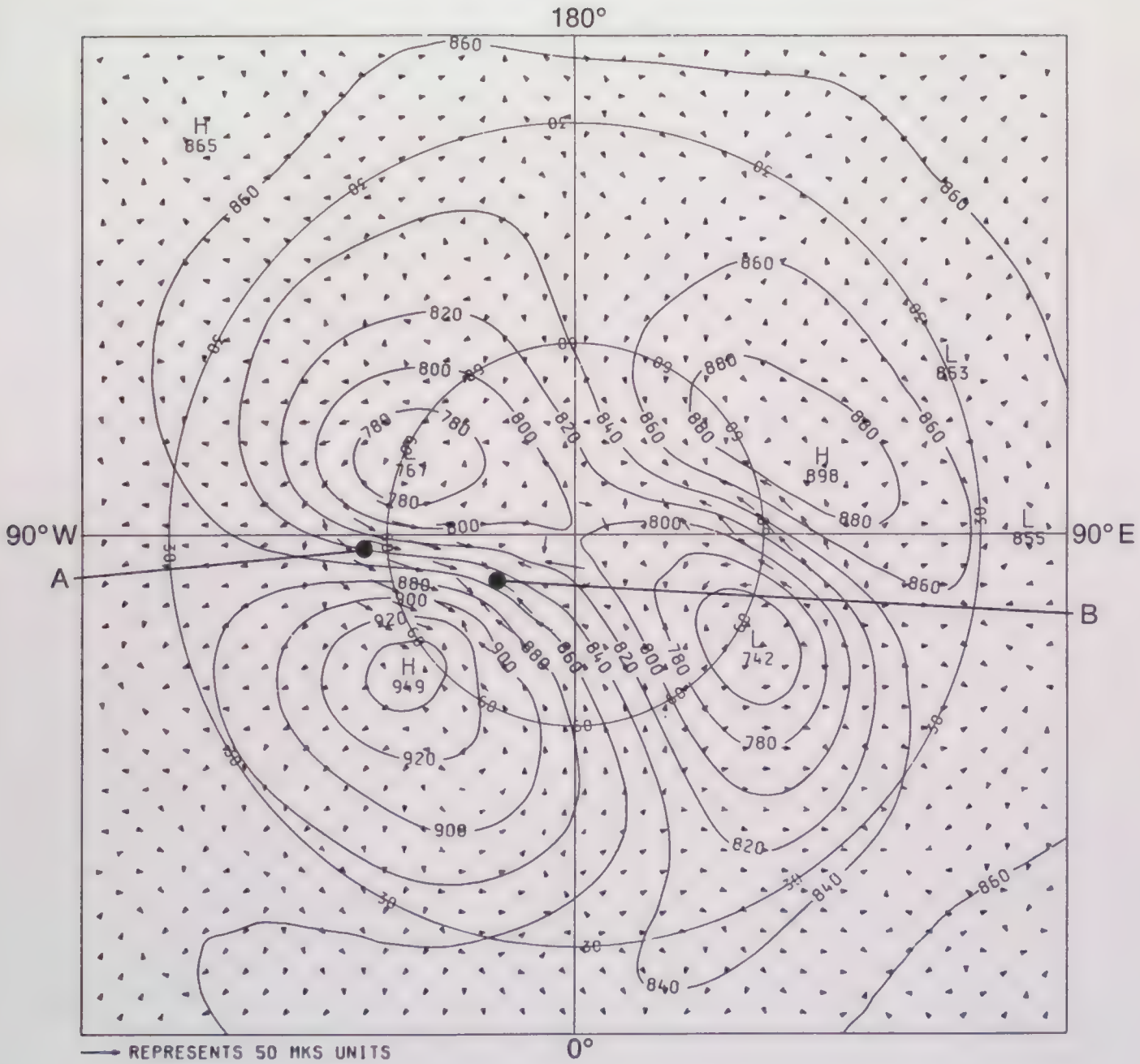
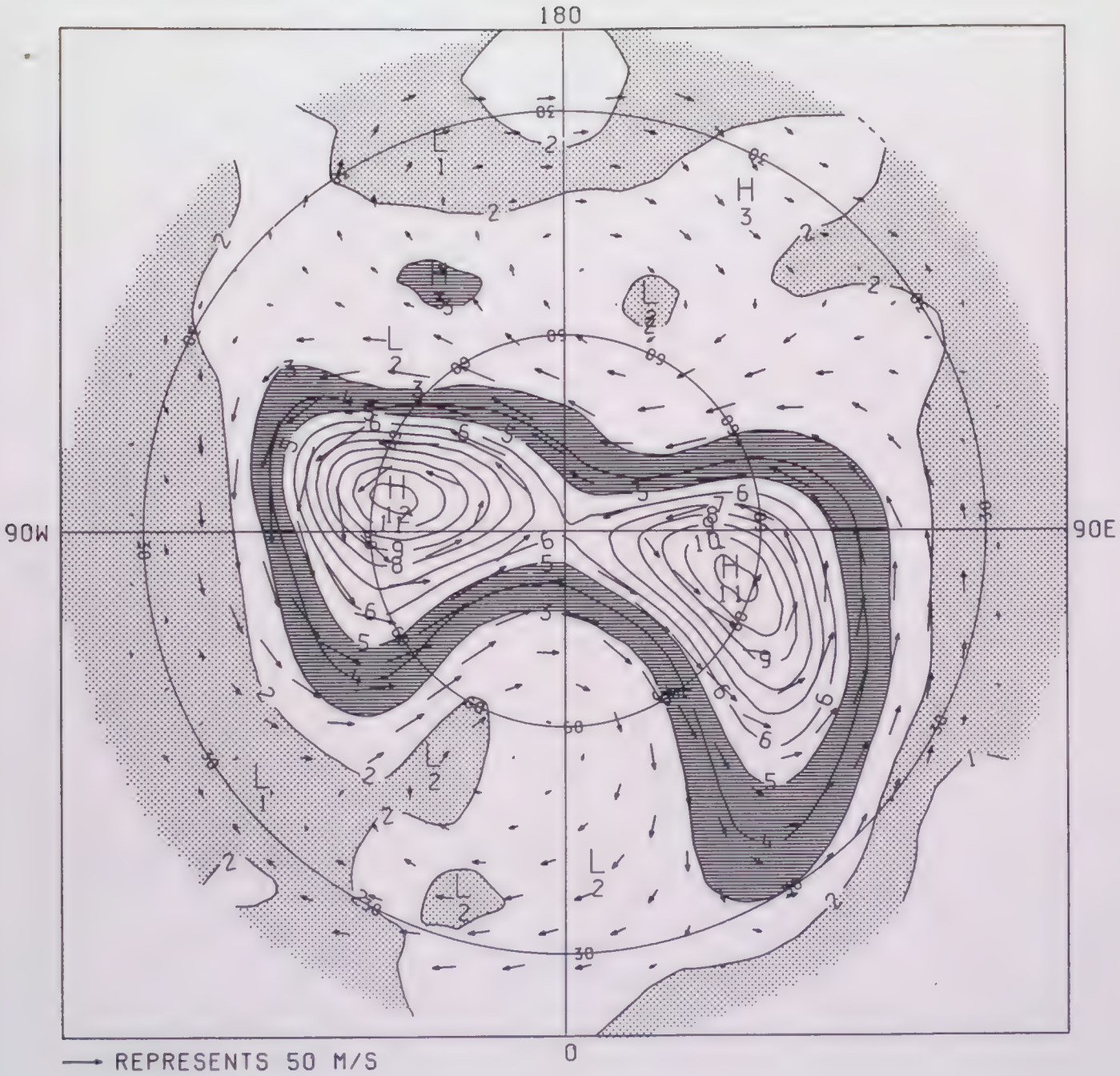

180°
90°W
90°E
0°
A
B
H
865
L
853
H
898
L
855
L
742
H
949
L
767
REPRESENTS 50 MKS UNITS

Figure 5



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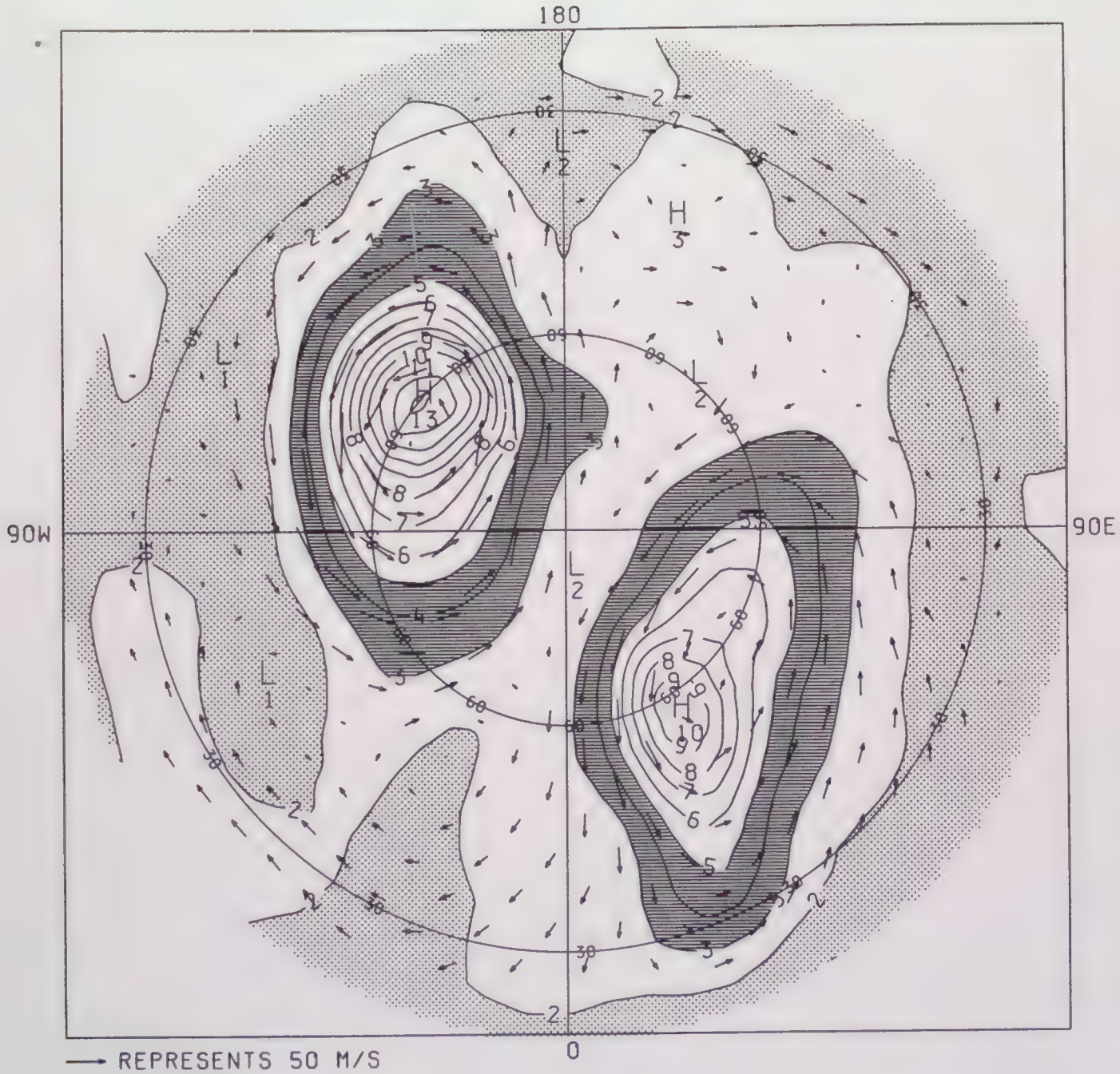

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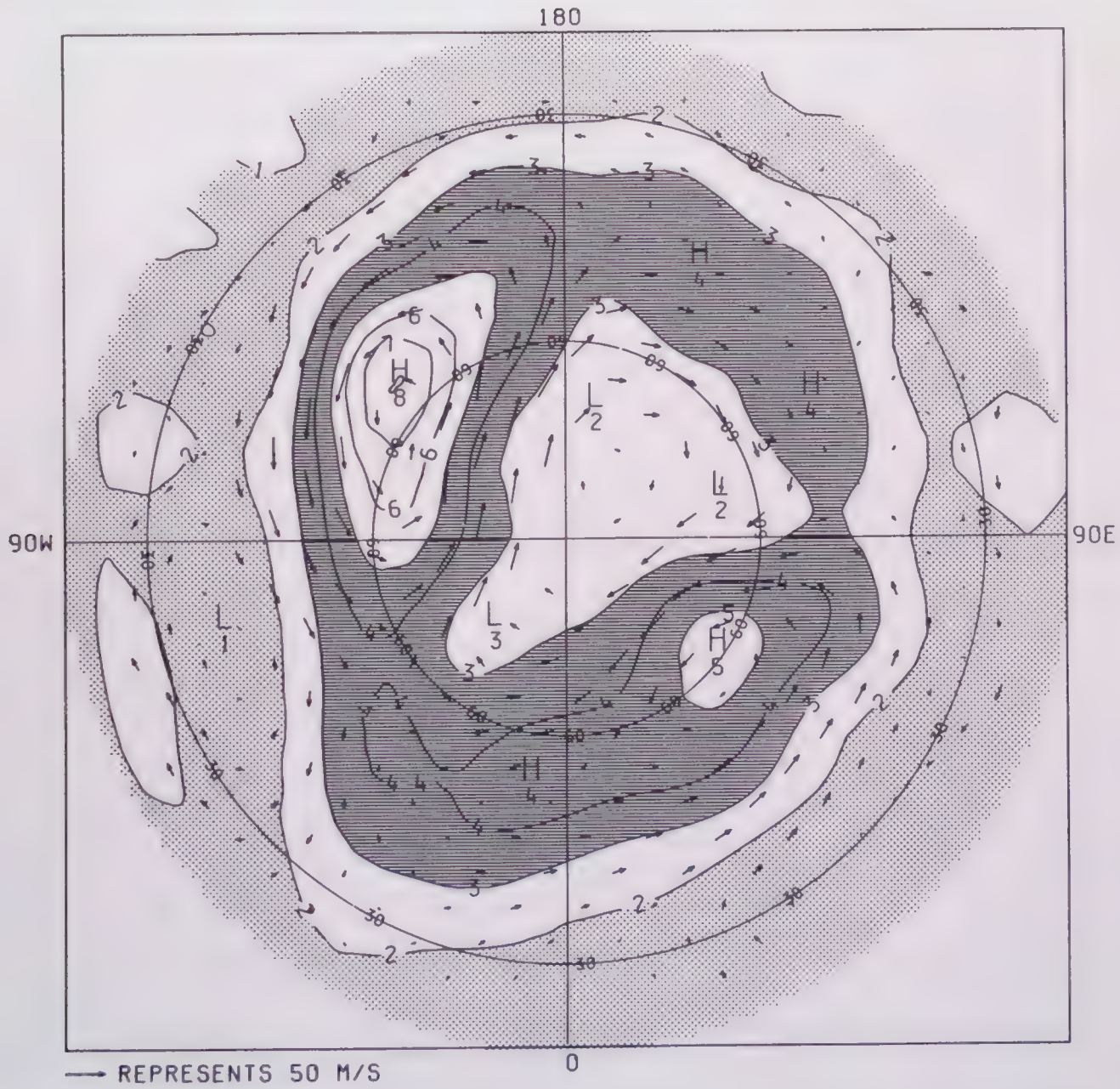


Figure 8

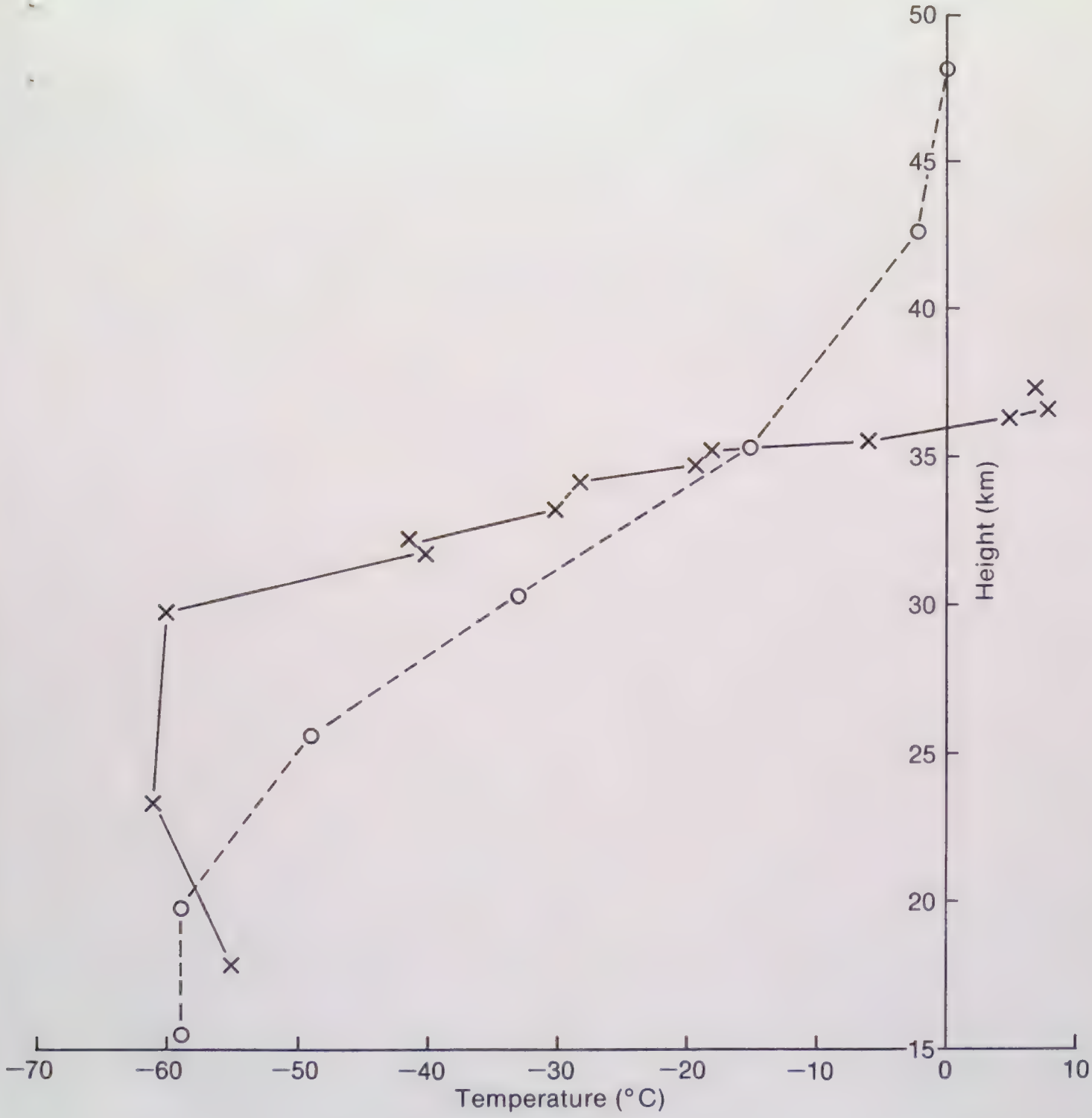

50
45
40
35
30
25
20
15
Height (km)
−70
−60
−50
−40
−30
−20
−10
0
10
Temperature (°C)

Figure 9

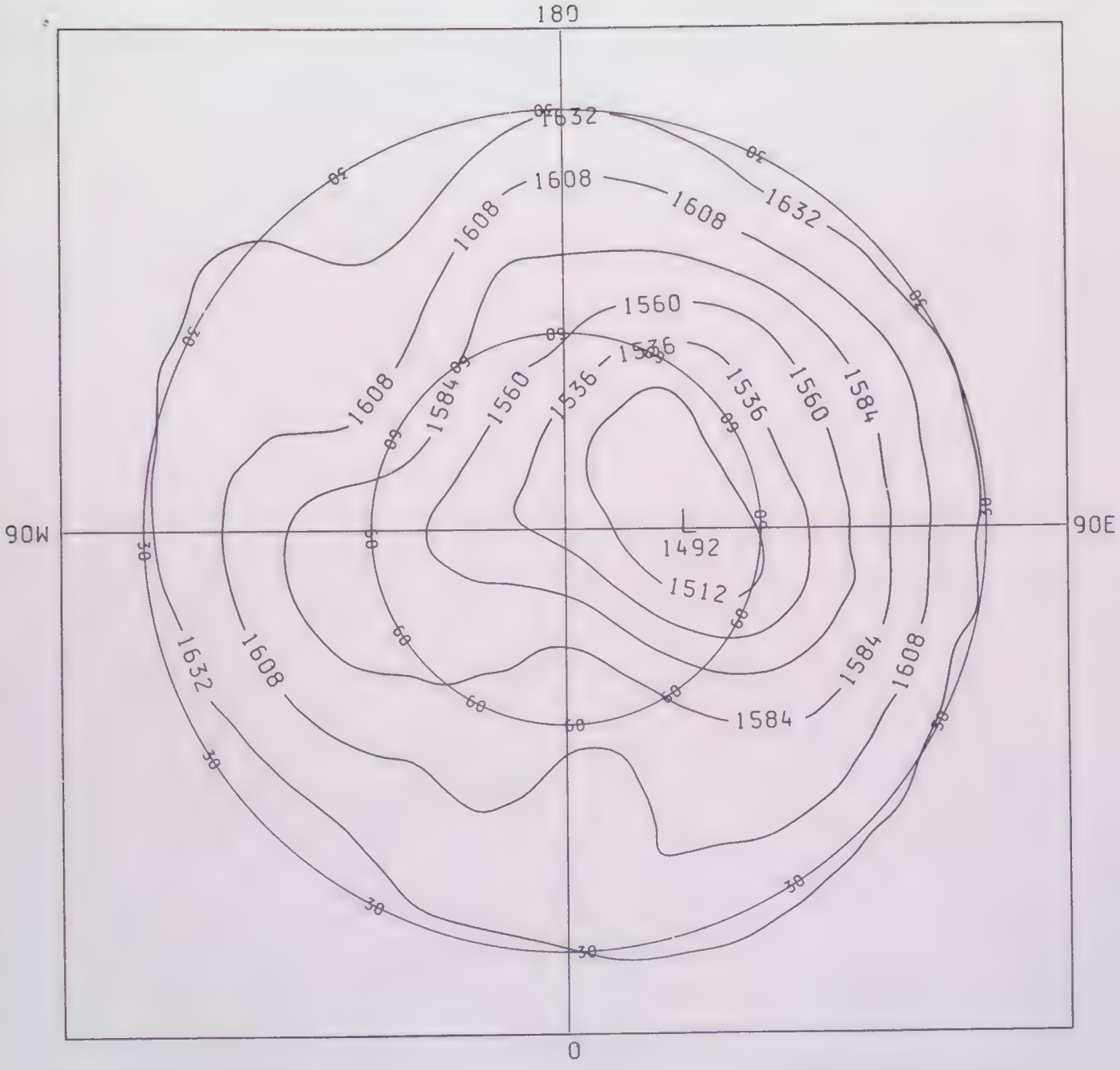


Figure 10

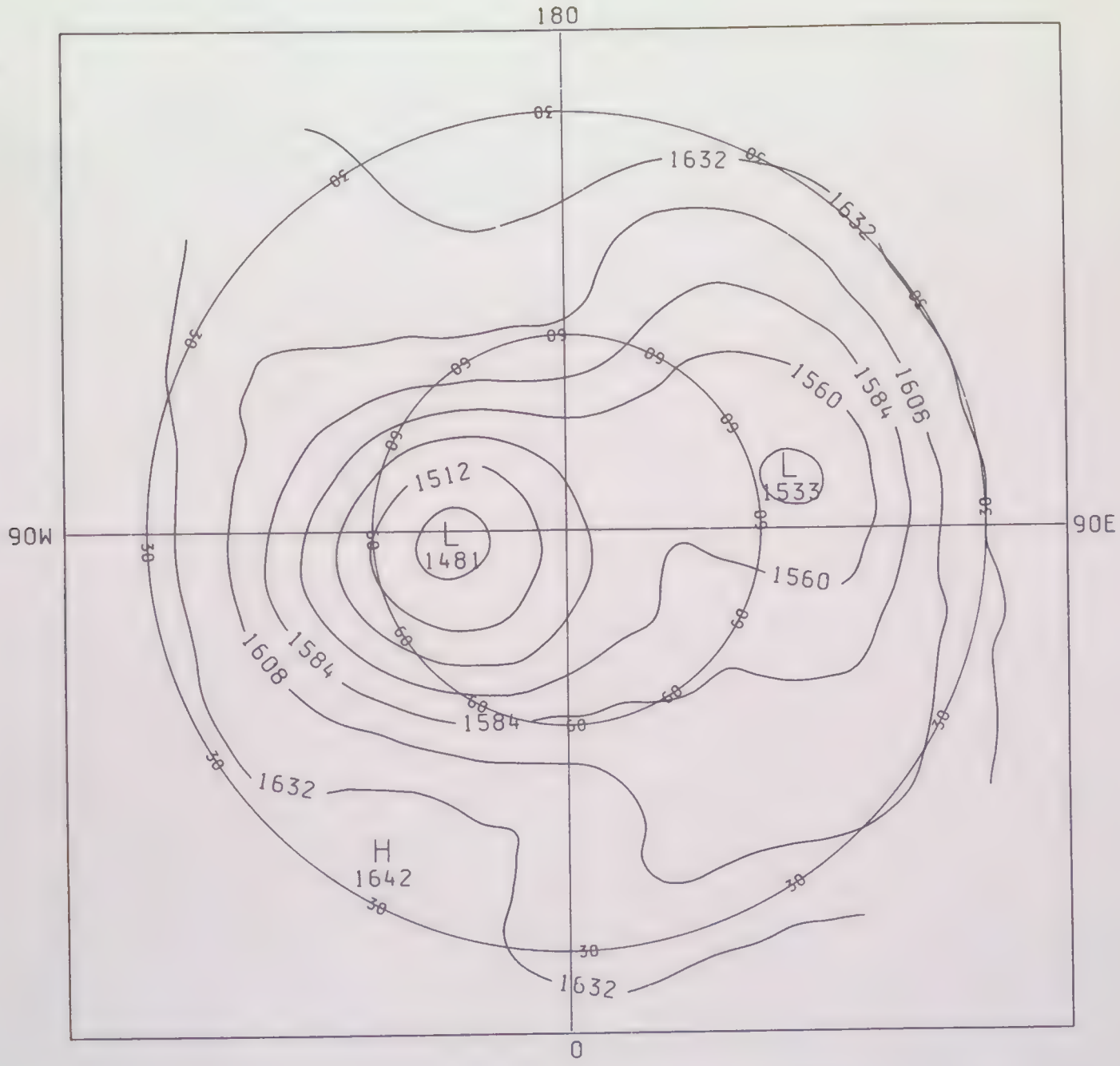
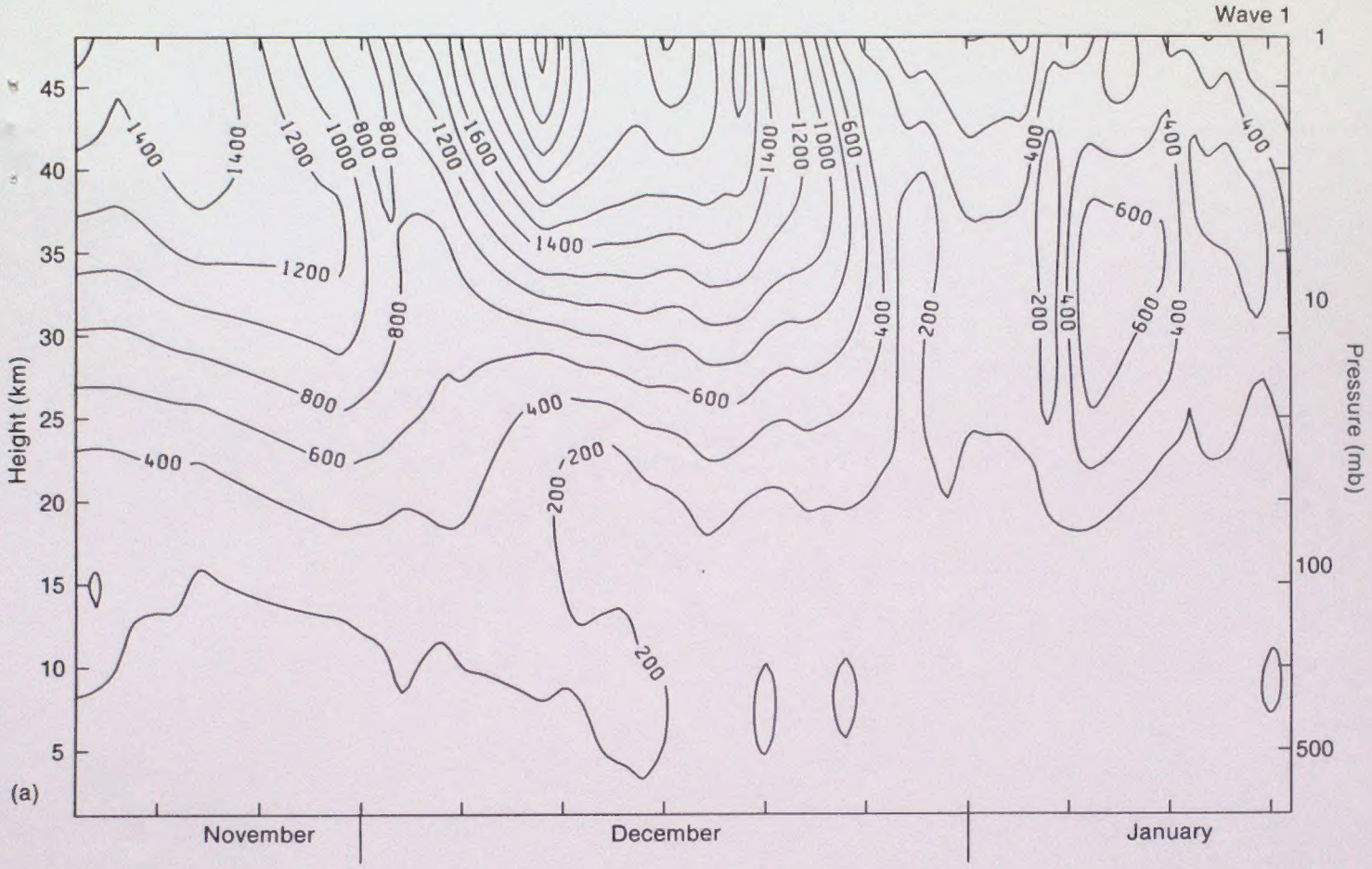
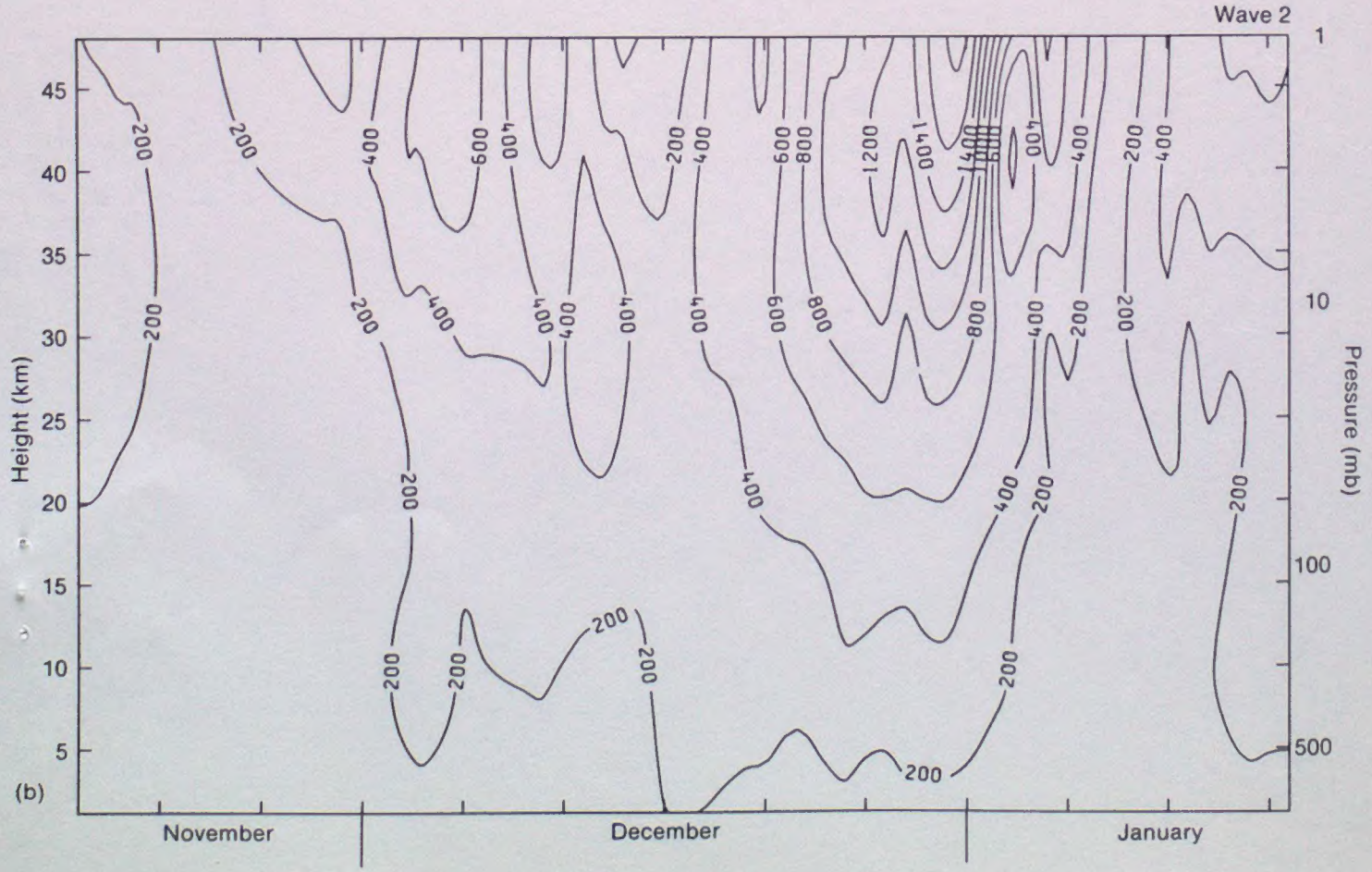

180
90W
90E
0
1632
1632
1560
1584
1606
1512
L
1533
L
1481
1560
1584
1608
1584
1632
H
1642
1632

Figure 11

Figure 12


Wave 1
Height (km)
Pressure (mb)
1
10
100
500
5
10
15
20
25
30
35
40
45
(a)
November
December
January


Wave 2
Height (km)
Pressure (mb)
1
10
100
500
5
10
15
20
25
30
35
40
45
(b)
November
December
January

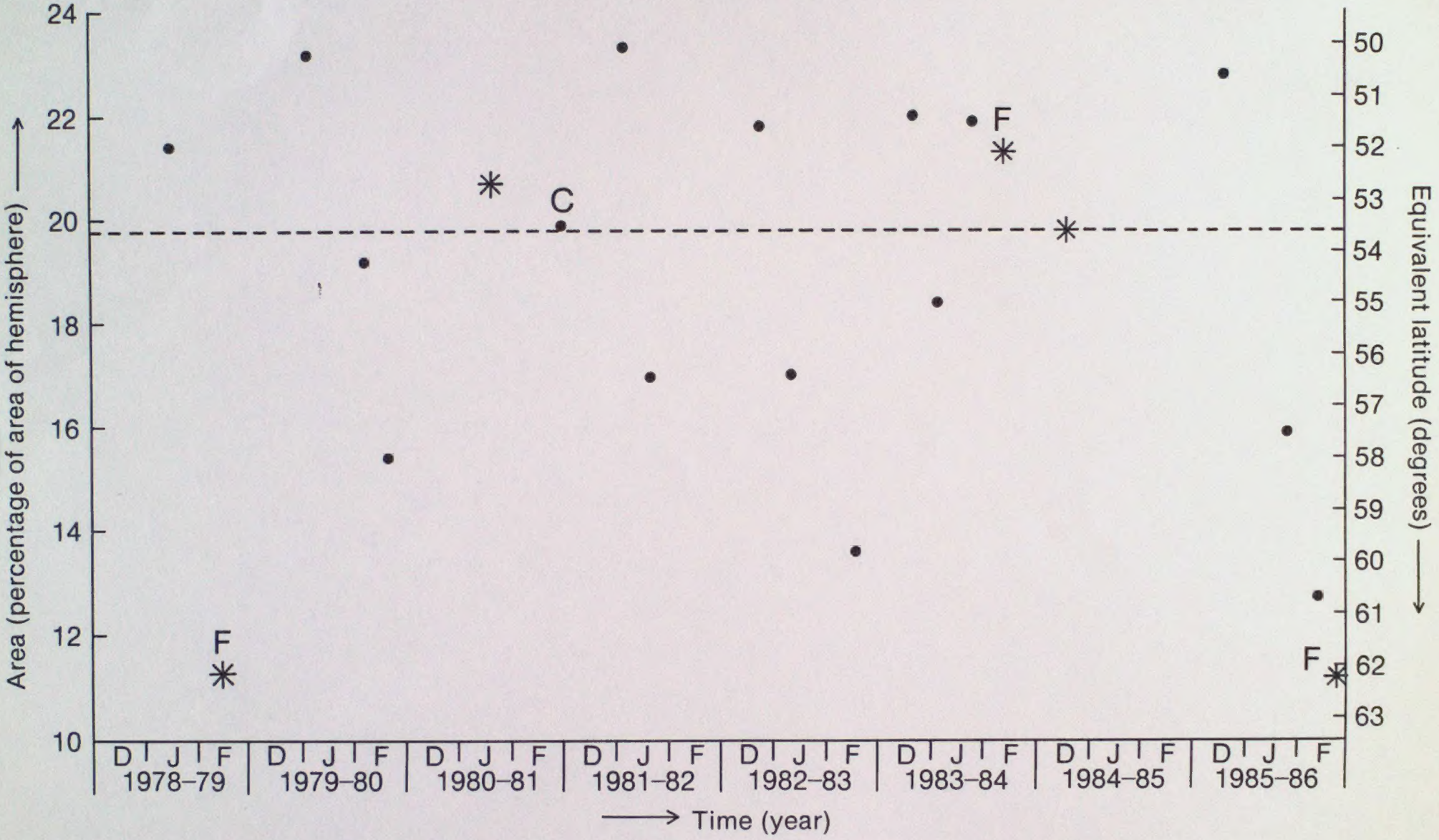

Area (percentage of area of hemisphere) ⟶
24
22
20
18
16
14
12
10
50
51
52
53
54
55
56
57
58
59
60
61
62
63
Equivalent latitude (degrees) ⟶
F
C
F
F
D J F
1978–79
1979–80
1980–81
1981–82
1982–83
1983–84
1984–85
1985–86
⟶ Time (year)

Figure 13

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